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# Subsurface structure in the vicinity of an intraplate earthquake swarm, central Arkansas

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ABSTRACT

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Over 40,000 events have been recorded in the Arkansas earthquake swarm since its inception in 1982. The earthquakes occur at depths between 3 and 6 km and cluster in a volume of about 25 km$^3$ beneath the easternmost Arkoma basin, near the town of Enola, Arkansas. A study of proprietary reflection seismic lines reveals that the earthquakes cluster within a graben formed in Mississippian time. This graben is part of a system of steeply dipping normal faults that trends ENE across the region. The regional strikes of the basement faults are not favorably oriented for activation under the regional stress regime. In the swarm area, however, these faults bend to form a 2.5-km long segment trending WNW. The small WNW striking segments are well oriented for left-lateral strike slip and focal mechanisms are consistent with this sense of slip. Additionally, a subset of the most accurately located earthquakes do appear to lie along a WNW trend. The length of the WNW trending fault segments is sufficient to have generated the largest of the swarm events. The reflection data reveal a loss of coherent reflectors within the swarm hypocentral volume. Reflectors above the graben have been uplifted about 30 m in post-Atokan (Pennsylvanian) time. Third order leveling surveys show 14.3 cm of uplift between 1961 and 1986 at a benchmark over the graben relative to a benchmark outside of the graben.

## Introduction

The Arkansas earthquake swarm, located about 50 km north of Little Rock near the town of Enola, Arkansas (Fig. 1), was first noted on January, 12, 1982 (Johnston, 1982). With over 40,000 earthquakes recorded since its inception, the earthquake swarm represents perhaps the largest number of events ever recorded in the central or eastern United States (Chiu et al., 1984). In this paper we use petroleum industry seismic reflection data to interpret the subsurface structure of the swarm source area and to correlate particular structures with the localization of swarm seismicity. We also review leveling data that indicate that present-day reactivation of these structures may be responsible for the swarm.

### *Geologic and seismotectonic setting*

The Arkansas earthquake swarm is located in the easternmost Arkoma basin of Arkansas, just north of the frontal thrust faults of the Ouachita transition zone (Fig. 1). A stratigraphic column for the study area is shown in Fig. 2 and comprises a Cambrian through Mississippian carbonate section and Pennsylvanian sandstones and shales with subordinate carbonates. This section overlies a postulated basal Cambrian clastic section and Precambrian granitic rocks (Burroughs, 1988). In the immediate area of the Arkansas swarm, the surface is underlain by about 3000 m

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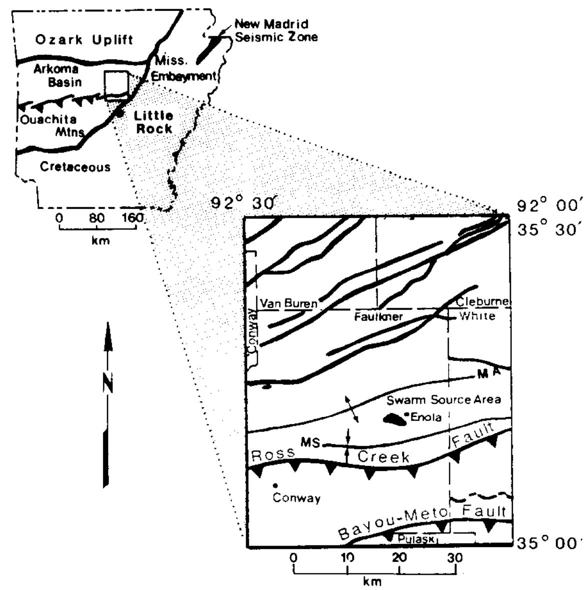


Fig. 1. Location map and generalized surface structure of the study area. Geologic provinces on upper map modified from Sutherland (1988). Lower map comprises the same area as Fig. 4. Outline of swarm area includes all but one earthquake relocated by Pujol et al. (1989). *MA* = Morrilton anticline, *MS* = Menifee syncline.

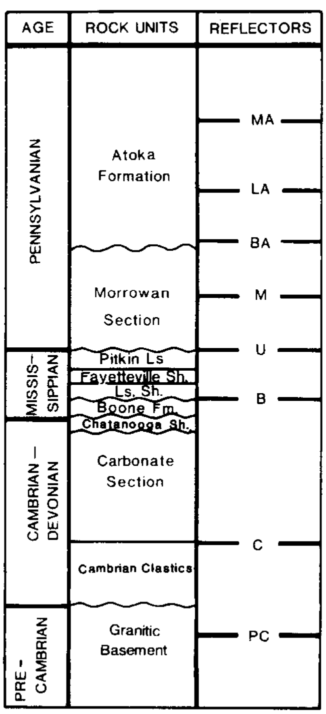


Fig. 2. Generalized subsurface stratigraphy (Caplan, 1954) in the Arkansas earthquake swarm region. Reflectors are those shown in Fig. 3.

of the Pennsylvanian Atoka Formation, which represents a thick clastic section that accumulated in the foreland basin as the Llanorian continent collided with North America (e.g., Houseknecht, 1986).

Surface structure in the swarm area consists of a series of sub-horizontal, upright, E-trending folds, bounded on the south by the Cadron anticline containing the Ross Creek thrust fault (Fig. 1). The swarm itself is occurring beneath the Menifee syncline.

Historically, seismic activity has been very low in central Arkansas (Chiu et al., 1984; Haar et al., 1984). Chiu et al. (1984) point out that prior to the swarm, central U.S. catalogs list only one nearby event, the $m_{\mathrm{bLg}} = 4.5$ Ferndale, Arkansas event of January 1, 1969, yet there were four events with $m_{\mathrm{bLg}}$ magnitudes between 4.0 and 4.5 during the first two months of the swarm (Pujol et al., 1989). Seismic activity continues at present, although at a considerably lower rate.

Johnston (1982) speculated that the Arkansas earthquake swarm was being caused by a shallow igneous intrusion, largely based on similarities between these events and those of the larger Matsushiro, Japan, swarm of 1965–1967. Chiu et al. (1984), noting dramatic changes in the $V_p/V_s$ ratio over periods as short as one day and the high density of earthquakes in a small volume, concluded that the rock in the swarm area must be highly fractured. They attributed the $V_p/V_s$ variations to opening and closing of fluid-filled cracks on a time scale of hours.

Pujol et al. (1989) have recently used a joint hypocentral determination technique to relocate events recorded in the Arkansas swarm area. They determine that the swarm earthquakes cluster in a volume of about 25 $km^3$ with depths between 3 and 6 km (Fig. 3). They also find that the hypocentral volume is characterized by seismic velocities averaging about 12–15% lower than the surrounding region, which could also be explained by the presence of fluids, such as water, natural gas, or magma.

Focal mechanisms of swarm events fairly consistently indicate right-lateral slip on north to north-northeast striking planes or left-lateral slip with a small thrust component on W to WNW

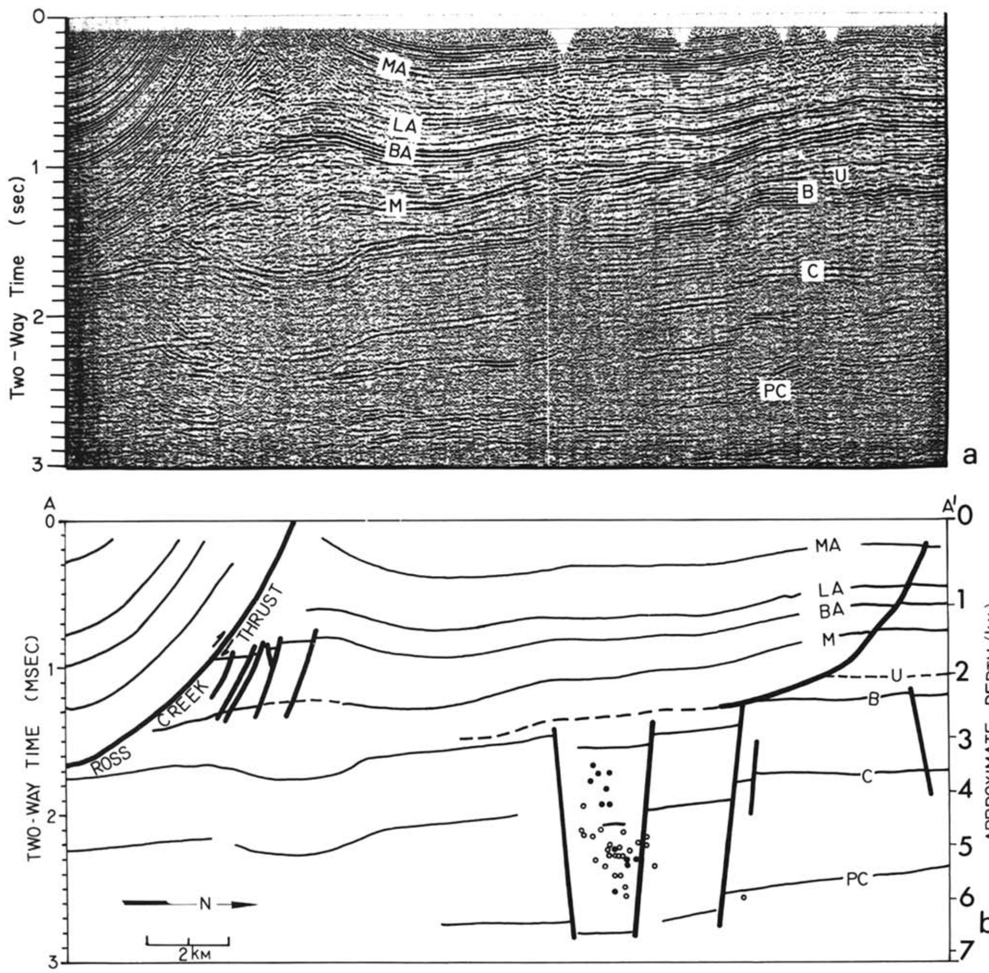


Fig. 3. Reflection seismic section $A$–$A'$ and interpretation. Location of the seismic section shown on Fig. 4. Earthquakes are those relocated by Pujol (1989) and projected parallel to the WNW trending faults. Open circles represent U.S.G.S. data from 1982; closed circles, Portable Array for Numerical Data Acquisition (PANDA) data from 1987. Depths are approximate and were determined using the velocity model of Chiu et al. (ms. in prep.). Reflectors: *MA* = middle Atoka; *LA* = lower Atoka; *BA* = Basal Atoka; *M* = Morrowan; *U* = pre-Morrowan unconformity; *B* = Boone Formation; *C* = top of Cambrian clastics; *PC* = Precambrian reflector.

striking planes, with compression directed east-northeast (Chiu et al., 1984; ms. in prep.).

*Data*

To gain an understanding of the structural geology at hypocentral depths, we undertook a study of 425 km of migrated reflection seismic lines in the eastern Arkoma basin of Arkansas. The area covered by these lines is about 1950 km$^2$, largely coincident with Faulkner County. The interpretation of the seismic data is controlled by surface geologic mapping and well logs (Burroughs, 1988). The regional results of this study are described in a companion paper (VanArsdale and Schweig, 1990).

The prominent reflectors traced on the seismic line (Fig. 3a) and illustrated on Fig. 3b are interpreted to be a middle Atoka sandstone (*MA*), a lower Atoka sandstone (*LA*), the basal Atoka sandstone (*BA*), a Morrowan age reflector (*M*), an unconformity that is post-Pitkin and pre-Morrowan in age (*U*), the Boone Formation (*B*), the top of the Cambrian clastic section (*C*), and a reflector in the Precambrian basement (*PC*) (VanArsdale and Schweig, 1990).

## Structure

Two different structural regimes are evident north of the Ross Creek thrust fault on the seismic reflection lines (Fig. 3), a deep basement regime and a near-surface regime. Both regimes are characterized by S-dipping normal faults. Normal faults in the near-surface regime are listric. The basement faults are high-angle, planar, and continuous from the Precambrian rocks into the Mississippian Pitkin Limestone. They are terminated by the pre-Morrowan unconformity, which marks the boundary between the two structural regimes. It is in the basement regime that the vast majority of the swarm events are occurring.

A two-way travel-time structure map of the Boone Formation reflector is shown in Fig. 4. The basement regime faults, which truncate the Boone reflector, are shown on the map, as is the location of the swarm area. Displacements of the Boone Formation reflector are calculated to range from 240 to 550 m. Over most of the map the faults generally trend N70°–75°E. In the immediate area of the swarm, however, the faults bend to form a 2.5-km-long segment trending about N80°–85°W. The epicenters, as recently relocated by Pujol et al. (1989), clearly map over a 2-km wide graben (Fig. 5). Projecting the earthquake hypocenters parallel to the trend of the graben faults onto Fig. 3a, which passes through

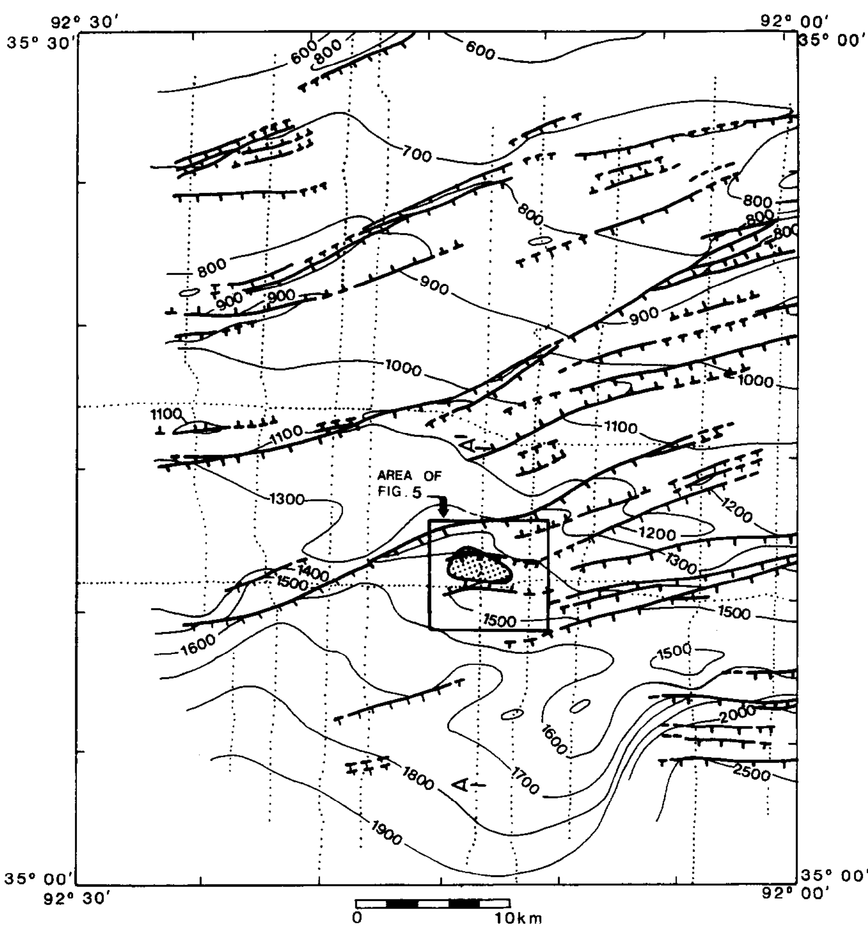


Fig. 4. Two-way travel-time structure map of the Boone Formation (reflector $B$ on Fig. 3). Contours in msec. Bold lines are normal faults, teeth on downthrown side. Dotted lines are the locations of the seismic reflection lines used in this study; $A$–$A'$ is the line shown on Fig. 3. Location of swarm shown in shaded pattern.

the source area, shows that the events actually occur within the graben, between the Boone and the basement reflectors.

The Boone Formation within the graben is down-dropped about 150 m. Above the Morrowan unconformity, however, the basal Atokan reflectors have been *uplifted* about 30 m, forming a broad, gentle anticline. Additionally, the Boone and deeper prominent reflectors become discontinuous within the hypocentral area (Fig. 3). This loss of coherent reflectors is also evident on an E–W seismic line (not shown here) passing just south of the swarm area.

The combination of a major bend in the basement fault regime, post-Atokan reactivation of basement faults, and the loss of coherent reflectors is unique to the swarm area within the region mapped. Any model of the seismicity of the Arkansas swarm area should consider these factors, as well as the localized low velocities in the swarm area determined by Pujol et al. (1989).

## Discussion

A relationship may exist among the seismicity of the Arkansas earthquake zone, the structural geometry at hypocentral depths, and the regional stress regime. The swarm area lies within the midcontinent stress province of Zoback and Zoback (1980). More recently Zoback and Zoback (1990) have expanded this province to include much of Canada and the eastern seaboard of the United States and renamed it the mid-plate province (see also Zoback et al., 1986). This is an area of fairly uniform compressive stress field with the maximum horizontal stress having an average orientation of east-northeast. Zoback and Zoback (1980) note that earthquake focal mechanisms throughout the province typically show both strike slip and thrust components, indicating that the intermediate and minimum principal stress magnitudes are similar in magnitude. In fact, Herrmann and Canas (1978) and O'Connell et al. (1982) pointed

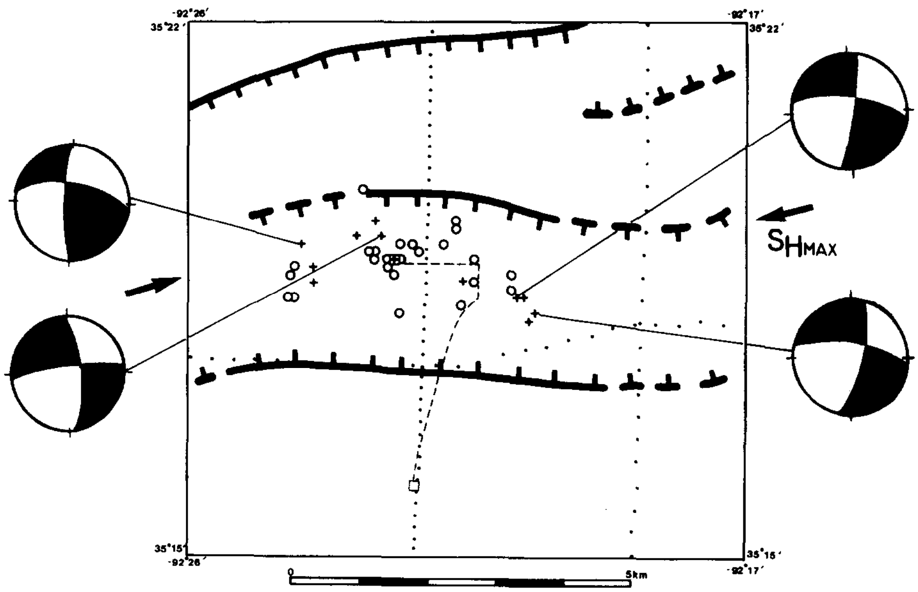


Fig. 5. Enlargement of a part of Fig. 4 showing faults in Boone Formation and swarm events. 1982 U.S.G.S. data shown as open circles, 1987 PANDA data as +'s. Dots represent points along seismic reflection lines. Representative lower hemisphere focal mechanisms from Chiu et al. (ms. in prep.) are shown for four events. Open squares and dashed line represent benchmarks and leveling line, respectively, discussed in text. $S_{Hmax}$ is approximate regional maximum horizontal stress direction from Zoback and Zoback (1989).

out that both types of events occur in different segments of the New Madrid seismic zone, 165 km northeast of Enola.

The strikes of the basement faults in the swarm region (N70°–75°E) would appear to be approximately parallel to the regional maximum compressive stress, an unfavorable orientation for producing high shear stress on these structures. The small WNW striking segments in the immediate vicinity of the swarm, however, are well-oriented for left-lateral strike-slip. In fact, most of the focal mechanisms for the swarm (Chiu et al., ms. in prep.) are compatible with left lateral slip on such a plane, generally with a small component of thrusting. Also, the earthquakes relocated by Pujol et al. (1989) do appear to lie along a WNW trend (Fig. 5), particularly the better-located events determined from the PANDA array maintained by the Center for Earthquake Research and Information (Chiu et al., ms. in prep.; Pujol et al., 1989). The 2.5 km WNW striking segment is clearly long enough to accommodate the largest of the swarm events. Nuttli (1983) has estimated fault dimensions (both length and width) of 2.1 km and 3.8 km for mid-plate earthquakes of $m_b$ 4.5 and 5.0, respectively. Thus it may be that it is the length of the WNW trending fault segment that has limited the size of the largest swarm events to $m_{bLg}$ 4.5.

The earthquakes do not lie directly on the bounding faults of the graben, but are mostly contained within it. Assuming that the earthquakes are accurately located, they may be occurring on parallel faults within the graben. The lack of continuous reflectors, however, makes this determination difficult.

One other piece of data relevant to modeling the Arkansas earthquake swarm is presented here. Haar et al. (1984) reported that a leveling line surveyed by E. Rowland, Arkansas State Surveyor, indicated 20 cm of uplift had occurred in the vicinity of the swarm area since emplacement of benchmarks in 1961. Burroughs (1988) conducted a third order leveling survey in 1986 to determine the change in vertical position of benchmark USGS 3 SAN 1961 relative to benchmark USC&GS Y 209 (Fig. 5). Benchmark USGS 3 SAN 1961 lies near the center of the swarm whereas USC&GS Y 209 lies outside the epicentral area. The survey revealed a vertical uplift of 14.3 cm of USGS 3 SAN 1961 relative to USC&GS Y 209 since 1961, corresponding to an average uplift rate of 0.57 cm per year. The error of closure on the loop was 8.0 mm. Thus the relative uplift far exceeds the possible survey error.

The recent uplift over the graben containing the swarm hypocenters could possibly be explained by the thrusting component of slip indicated by the focal mechanisms. In this model the block between the graben-bounding faults is being squeezed up due to a component of compression across the graben, similar to uplift between strands of the San Andreas Fault northeast of the Salton Sea, California, described by Sylvester and Smith (1976). We do not, however, know the age of most of this uplift, except that it is post-Atokan in age and that some uplift has occurred in the last 25 years.

The lack of continuous reflectors in the hypocentral area could be explained by a number of factors including poor data or data processing, steeply dipping structures, and intense fracturing in the graben. We favor the latter explanation for the following reasons. First, although the N–S seismic line (Fig. 3) does have a 1 km-long data gap above the swarm area that could account for discontinuous reflectors below, there are reflectors above and below the hypocentral area that *are* continuous. However, an E–W line passing just south of the swarm (Fig. 4) with no data gap also shows the lack of continuous reflectors. Also, there are other data gaps of similar length north of the swarm area that do not have a major effect on reflectors at hypocentral depth. McCarthy and Thompson (1988) point out that seismic reflection lines commonly show a transparent region bracketing major fault zones that may be due, in part, to steeply dipping structures that are difficult to resolve. In the region surrounding the Arkansas swarm area, however, there are many steeply dipping faults without transparent zones. Also, the faults bounding the graben that contains the swarm hypocenters are well imaged; the lack of reflectors is within the graben itself. If, as suggested above, the restraining bend in the graben has been undergoing transpressional strike-slip faulting, intense fracturing would be expected. This is, in fact, supported by the strong low-velocity anomaly found in the epicentral area by Pujol et al. (1989), which could be the result of fluid-filled cracks. The low-velocity anomaly is not constrained in depth.

## Conclusions

The Arkansas earthquake swarm is apparently occurring in a highly fractured fault zone contained within a bend in a graben formed in Mississippian time. Although this study does not explain why an earthquake swarm is occurring, or why the seismic strain release was of a swarm character rather than a normal mainshock–aftershock sequence, it does show that there is a combination of characteristics within the swarm area that may favor seismogenesis. These characteristics include a bend in the basement regime normal faults from ENE to WNW in the vicinity of the swarm, loss of continuity in otherwise-strong reflectors, uplift of the Atokan section above the hypocentral area in seismic reflection data as well as in recent leveling surveys, lower seismic velocities in the swarm area relative to the surroundings, and changes in the $V_p/V_s$ ratio over short time periods. Fluid migration would appear to be playing a role in the velocity characteristics of the swarm area. Whether the fluid is magma, as speculated by Johnston (1982), water (Costain et al., 1987; Pujol et al., 1989), or natural gas is not known, although the regional geology suggests that migrating magma is unlikely. As pointed out by Pujol et al. (1989), additional data are necessary to determine the cause of the velocity anomaly.

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